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# Rayleigh-wave phase velocities in French Polynesia

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**Summary.** Rayleigh-wave phase velocities are investigated in the period range 17–100 s by the two-station method over several paths covering most of French Polynesia. Our results confirm the validity of theoretical models obtained through regionalization of data pertaining to longer paths. They also exhibit a 2–3.5 per cent anisotropy, with the axis of maximum velocity oriented in the direction of spreading of the plate. Part of this anisotropy is, however, due to the presence of the Tuamotu archipelago; when this is removed, the remaining anisotropy (about 1.5 per cent) correlates with the present direction of spreading, indicating that a relaxation of the anisotropy has taken place since the East Pacific ridge jump. Finally, the presence of the Tuamotu Islands explains anomalous waveshapes for surface waves travelling in their vicinity, due to multipathing through their faster structure.

## 1 Introduction

Because of sparse station coverage, most of our knowledge of the upper mantle structure beneath oceanic areas has been obtained from the study of seismic surface wave dispersion. Early investigations (e.g. Aki & Press 1961) attempted to use a single oceanic model but it soon became apparent that the ageing of the plates away from oceanic ridges creates lateral heterogeneity inside the oceanic upper mantle, whose properties must then be described in terms of several models, relative to portions of the ocean of different lithospheric age. However, due to the loose network of oceanic stations, it is then difficult, if not impossible, to isolate paths sampling areas of the ocean of homogeneous age, and 'pure-path' models, corresponding to individual regions of given lithospheric age, have been obtained through regionalization of a large number of heterogeneous paths, sampling these regions more or less at random. This procedure has been applied extensively in the Pacific Ocean (Kausel, Leeds & Knopoff 1974; Leeds 1975; Forsyth 1975; Yu & Mitchell 1979). Results have generally indicated an increase in phase and group velocities with the age of the plate, a property often interpreted as a thickening of the lithosphere at the expense of the asthenosphere, as the plate cools down away from the ridge (Leeds, Kausel & Knopoff 1974; Yoshii 1975; Yu & Mitchell 1979). Additionally, detailed seismic studies by Forsyth (1975), Schlue & Knopoff (1977) and Yu & Mitchell (1979) have evidenced azimuthal as well as polarization anisotropy in the propagation of Love and Rayleigh waves. The faster velocities are observed in the direction of spreading of the plate, and *SH* velocities appear to be greater than their *SV* counterparts.

Most of these results were, however, obtained from the single-station method (Brune, Nafe & Oliver 1960), which requires accurate knowledge of the earthquake's focal mechanism, depth and epicentral location, in order to retrieve the phase velocity information. In particular, Mitchell & Yu (1980) have recently revised some of their 1979 models, partly because of the reassessment of two epicentral locations in their earlier study. It therefore seems desirable to obtain direct confirmation of the validity of the 'pure-path' models obtained through regionalization, by a local study of the surface wave dispersion, confined to the interior of one region of homogeneous age. In the present paper, we use data from a five-station vertical long-period array in French Polynesia to perform independent measurements of Rayleigh-wave phase velocities, using the two-station method along paths oriented at various azimuths. Our conclusions indicate the adequacy in this area of the Pacific of the Mitchell & Yu (1980) models, and confirm the presence of azimuthal anisotropy well correlated with the direction of motion of the plate. However, part of this azimuthal anisotropy is found to be due to the large Tuamotu island archipelago, leading to multipathing of the surface waves travelling parallel to it, but outside the island chain.

## 2 Data

The French Polynesian long-period network is mapped on Fig. 1, and station coordinates are given in Table 1. Until very recently (1979), only PPT was equipped with horizontal instruments; therefore, our study will be limited to Rayleigh waves. The characteristics of the instruments, and the various frequency bands used for recording have been described elsewhere (Talandier 1972).

This network is located in an area of the Pacific whose age is only approximately known. Magnetic anomalies (Herron 1972) suggest lithospheric ages of 42 Myr around the Gambier

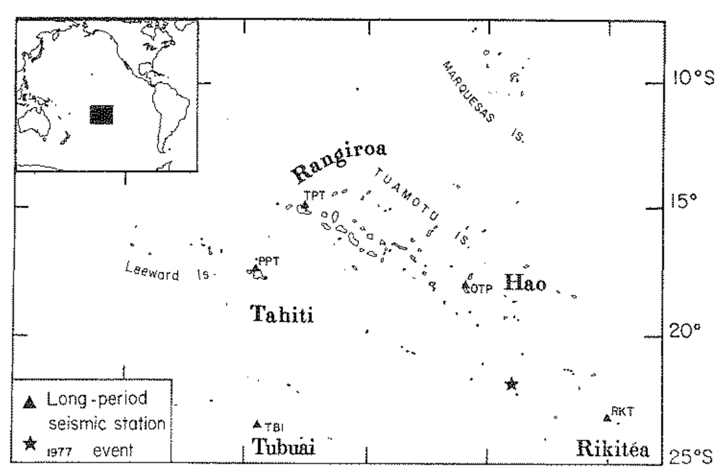


Figure 1. Map of the long-period seismic network in French Polynesia. Also shown is the epicentre of the 1977 March 19 event used in Section 3.

**Table 1.** Long-period stations of the French Polynesian network.

| Code | Name | Island | Archipelago | Latitude (° S) | Longitude (° W) | Geological setting |
|---|---|---|---|---|---|---|
| PPT | Papeete | Tahiti | Society | 17.569 | 149.576 | Volcanic |
| TPT | Tiputa | Rangiroa | Tuamotu | 14.984 | 147.620 | Atoll |
| OTP | Otepa | Hao | Tuamotu | 18.167 | 140.857 | Atoll |
| RKT | Rikitéa | Mangaréva | Gambier | 23.120 | 134.973 | Volcanic |
| TBI | Tubuai | Tubuai | Austral | 23.349 | 149.461 | Volcanic |

Islands, and 68 Myr at the western end of the Tuamotus. The islands in the area are the expression or the remnants of more recent volcanism; the youngest ones have been accurately dated (Tahiti: 2 Myr and less; Gambier: 6 Myr), but the oldest ones subsist only as coral reefs (e.g. Rangiroa), and only loose bounds exist on their ages: DSDP site 318, north-east of Rangiroa yielded ages of 49 Myr for the sediments capping the unreached volcanic edifice, suggesting that the age of this portion of the Tuamotu chain is grossly between 50 and 70 Myr (Jackson & Schlanger 1976). Also, the Tuamotu Islands are unique among Pacific island chains, in the transverse dimension of the archipelago (up to 300 km), and for the shallow waters separating the various islands, suggesting that their genesis may have been more complex than a simple hotspot model would predict. Accordingly, we will compare the results in the present study with models for Eocene to Cretaceous lithosphere, such as Mitchell & Yu's (1980) models 2 and 3, but will keep in mind that the Tuamotu chain may alter the local structure substantially.

Table 2 lists events used in this study, for which a good alignment (usually within 2° of azimuth at the source) was obtained for two distant stations in the network, allowing the use of the two-station method. The distribution of the events is shown on Fig. 2. We excluded any alignments involving the path RKT–PPT, since multipathing phenomena are

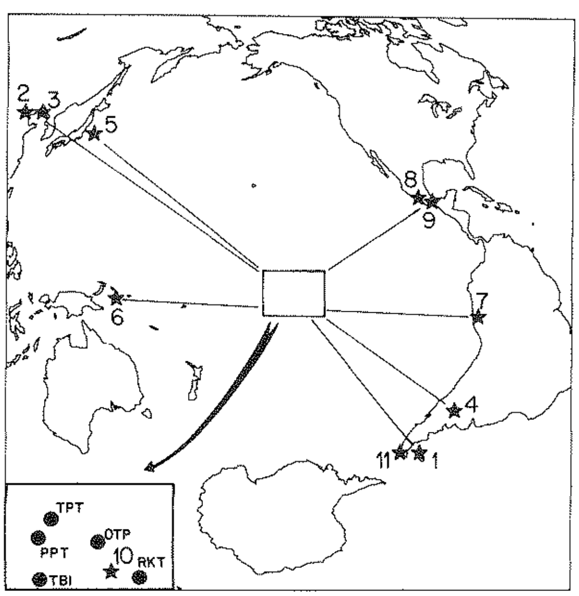


**Figure 2.** Map of the epicentres used in the present study. The projection is azimuthal equidistant, with its pole at the centre of the network. The inset at lower left outlines the network and the location of event 10. Stars each identify one event; circles are stations.

**Table 2.** Events and records used in the present study.

| No. | Date | Origin time (GMT) | Epicentre | | Depth (km) | $M_s$ | Stations used code | dist. (°) | azim. (°) | code | dist. (°) | azim. (°) | Remarks |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 1975 Dec. 29 | 03:39:43.0 | 56.80° S | 68.42° W | 18 | 6.5 | RKT | 58.09 | 276 | TPT | 71.63 | 271 | Southern Chile |
| 2 | 1976 Jul. 27 | 19:42:54.6 | 39.57° N | 117.98° E | 23 | 7.9 | RKT | 117.20 | 98 | TPT | 102.77 | 99 | Tangshan |
| 3 | 1976 Jul. 28 | 10:45:35.2 | 39.66° N | 118.40° E | 26 | 7.4 | RKT | 116.90 | 98 | TPT | 102.47 | 99 | Tangshan aftershock |
| 4 | 1977 Nov. 23 | 09:26:24.7 | 31.03° S | 67.77° W | 13 | 7.4 | RKT | 59.53 | 260 | TPT | 73.82 | 262 | San Juan, Argentina |
| 5 | 1978 Jun. 12 | 08:14:26.4 | 38.19° N | 142.03° E | 44 | 7.7 | RKT | 98.82 | 112 | TPT | 84.45 | 114 | Sendai |
| 6 | 1977 Jul. 29 | 11:15:45.3 | 8.03° S | 155.54° E | 33 | 7.2 | RKT | 68.05 | 112 | TBI | 54.77 | 113 | Solomon Islands |
| 7 | 1979 Feb. 16 | 10:08:54.5 | 16.43° S | 72.49° W | 53 | 6.7 | RKT | 58.78 | 253 | TBI | 71.93 | 250 | Peru |
| 8 | 1978 Nov. 29 | 19:52:47.6 | 16.01° N | 96.59° W | 20 | 7.8 | OTP | 55.32 | 234 | TBI | 64.88 | 234 | Oaxaca, Mexico |
| 9 | 1978 Feb. 22 | 06:07:37.0 | 14.17° N | 91.34° W | 100 | 6¼ | OTP | 58.50 | 238 | TBI | 68.04 | 237 | Guatemala |
| 10 | 1977 Mar. 19 | 23:00:58.3 | 21.93° S | 138.96° W | 0 | 5.9* | PPT | 10.90 | 292 | RKT | 3.87 | 109 | See Section 3 |
| 11 | 1970 Jun. 14 | 00:00:11.3 | 51.95° S | 73.84° W | 10 | 6.5 | PPT | 67.57 | 270 | TPT | 68.33 | 273 | Chile; see Section 4 |

* Body-wave magnitude $m_b$.

frequently observed along this path (e.g. for event 11). This will be discussed in detail in Section 4. Also included in Table 2 is a local 1977 event (No. 10), located south of the Tuamotu chain, whose data will be discussed in Section 3. In the present section, we will deal only with data from the first nine events, pertaining to the paths RKT–TPT, RKT–TBI and OTP–TBI.

Seismograms were digitized at a sampling of 1 s, and the phase velocity retrieved from their Fourier spectrum by the classic formula (Toksöz & Ben-Menahem 1963):

$$C(T) = \frac{\Delta_2 - \Delta_1}{t_2 - t_1 + T[(\phi_1 - \phi_2)/2\pi + N]}. \tag{1}$$

Here, the $\Delta$s are epicentral distances, the $\phi$s the spectral phases at period $T$, the $t$s the starting times of the digitization windows, and $N$ a suitable integer.

In comparing observations of phase velocities with values obtained from theoretical models, it is necessary to take into account the influence of the Earth's anelasticity (Liu, Anderson & Kanamori 1976). Since such corrections may depend on the choice of a reference frequency, and on the still poorly known behaviour of attenuation with frequency, we will choose to compare our data only to regionalized models of observed data, such as those proposed by Mitchell & Yu (1980), which were uncorrected for anelasticity. We will not compare our observations to theoretical dispersion curves obtained from purely elastic Earth models, such as that of Leeds *et al.* (1974). In this way, we do not need any corrections to the values derived through equation (1).

The uncertainties involved in the two-station method are difficult to assess; in this respect, this method, which eliminates source-dependent parameters, is more precise than the single-station method, and permits the use of relatively shorter paths. The remaining uncertainties come mostly from digitizing inaccuracies. We estimate an uncertainty of twice the digitizing interval, equivalent to 0.02 km s$^{-1}$ on $C$, for paths on the order of 1500 km. This estimate is comparable to the standard deviation of our results for different events along the same path, and also to the precision claimed in other surface wave studies (Forsyth 1975; Yu & Mitchell 1979).

Results are listed in Table 3, and sketched on Figs 3, 4 and 5. On each of the figures, the continuous lines represent the regionalized models No. 3 (top line) and No. 2 (bottom line) of Mitchell & Yu (1980). Their model 3 pertains to lithosphere aged 50–100 Myr, and their model 2 to an age of 20–50 Myr. If we estimate that the average age of lithosphere involved in the present study is 55 Myr, we must compare our results to values intermediate between their models 2 and 3. It is immediately apparent that azimuthal heterogeneity is present in our data, but also that the average dispersion of Rayleigh waves over French Polynesia in the period range 17–100 s is correctly described by Mitchell & Yu's regionalized models. Given the excellent agreement between their regionalized data and the theoretical dispersion curves they obtained from regionalized shear velocity models (after allowing for anelasticity), we conclude that these shear models correctly predict the average dispersion in this part of the Pacific, where paths are of homogeneous age.

The azimuthal heterogeneity evidenced by our data ranges from 2 per cent at 17 s to about 3.5 per cent at 73 s. This heterogeneity may be structural, or it may be due to anisotropy. If the latter, such an increase in the effect of anisotropy at longer periods would be in agreement with Forsyth's (1975) findings in the Nazca plate. Also, the fastest velocities are found along the axis of the Tuamotu chain, which corresponds to the direction of spreading; the slowest velocities along the path OTP–TBI, most transverse to it, with the east–west path RKT–TBI showing intermediate values, in excellent agreement also with Talandier & Bouchon's (1979) results on $P_n$ velocities in the same area. If we then choose to

**Table 3.** Rayleigh-wave phase velocities from the two-station method.

| Period (s) | Phase velocity ($km\ s^{-1}$) along: RKT–TPT | RKT–TBI | OTP–TBI |
|---|---|---|---|
| 102.40 | 4.071 | | |
| 93.09 | 4.069 | | |
| 85.33 | 4.082 | | |
| 78.77 | 4.032 | | |
| 73.14 | 4.002 | | 3.865 |
| 68.27 | 4.007 | | 3.874 |
| 64.00 | 4.026 | | 3.890 |
| 60.24 | 4.032 | | 3.849 |
| 56.89 | 4.029 | | 3.866 |
| 53.89 | 4.037 | | 3.886 |
| 51.20 | 4.041 | | 3.893 |
| 48.76 | 4.050 | 4.019 | 3.877 |
| 46.55 | 4.078 | 3.989 | 3.867 |
| 44.52 | 4.057 | 3.957 | 3.857 |
| 42.67 | 4.055 | 3.986 | 3.872 |
| 40.96 | 4.073 | 3.988 | 3.886 |
| 39.38 | 4.085 | 4.007 | 3.900 |
| 37.93 | 4.090 | 3.993 | 3.928 |
| 36.57 | 4.083 | 4.009 | 3.930 |
| 35.31 | 4.088 | 4.003 | 3.898 |
| 34.13 | 4.105 | 4.013 | 3.905 |
| 33.03 | 4.104 | 4.007 | 3.904 |
| 32.00 | 4.106 | 3.976 | 3.898 |
| 31.03 | 4.101 | 3.977 | 3.912 |
| 30.12 | 4.090 | 3.978 | 3.915 |
| 29.26 | 4.077 | 3.963 | 3.917 |
| 28.44 | 4.074 | 3.953 | 3.922 |
| 27.68 | 4.070 | 3.945 | 3.927 |
| 26.95 | | 3.922 | 3.933 |
| 26.26 | 4.095 | 3.917 | 3.928 |
| 25.60 | 4.102 | 3.926 | 3.919 |
| 24.98 | 4.087 | | 3.932 |
| 24.38 | 4.075 | | 3.936 |
| 23.81 | 4.056 | | 3.920 |
| 23.27 | 4.054 | | 3.922 |
| 22.76 | 4.043 | | 3.920 |
| 22.26 | 4.044 | | 3.917 |
| 21.79 | 4.031 | | 3.925 |
| 21.33 | 4.037 | | 3.921 |
| 20.90 | 4.048 | | 3.915 |
| 20.48 | 4.021 | | 3.922 |
| 20.08 | 4.026 | | 3.909 |
| 19.69 | 4.038 | | 3.909 |
| 19.32 | 4.014 | | 3.900 |
| 18.96 | 4.013 | | 3.889 |
| 18.62 | 4.002 | | 3.889 |
| 18.29 | 3.987 | | 3.889 |
| 17.96 | 3.975 | | 3.881 |
| 17.66 | 3.967 | | 3.875 |
| 17.36 | 3.961 | | 3.874 |

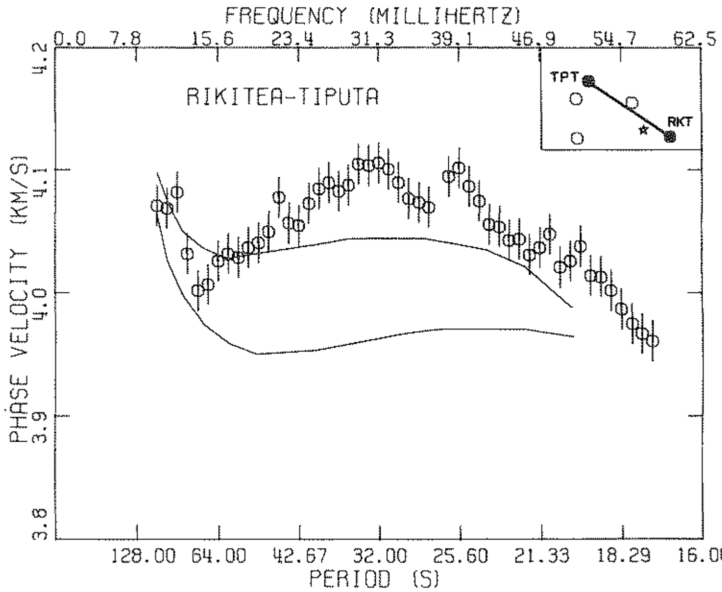


**Figure 3.** Rayleigh-wave phase velocities along the path RKT–TPT. The two solid lines represent Mitchell & Yu's (1980) regionalized models 3 (top) and 2 (bottom). Open circles (with error bars) represent our data, obtained in the present study. The scale is linear with frequency; for convenience, the lower scale shows periods. The inset at upper right sketches the position of the path studied in the network (see Figs 1 and 2).

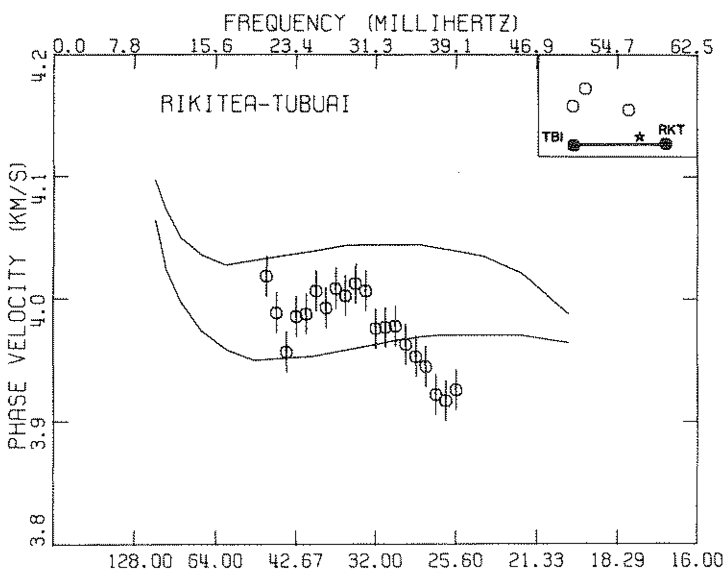


**Figure 4.** Same as Fig. 3 for the path RKT–TBI.

model the heterogeneity between the two extreme paths RKT–TPT and OTP–TBI as an azimuthal anisotropy of the form

$$C(T, \phi) = C(T)\,[1 + a(T) \cdot \cos 2(\phi - \phi_{\mathrm{pl}})], \tag{2}$$

where $\phi_{\mathrm{pl}}$ is the azimuth of motion of the plate (293°), and $\phi$ the azimuth of propagation

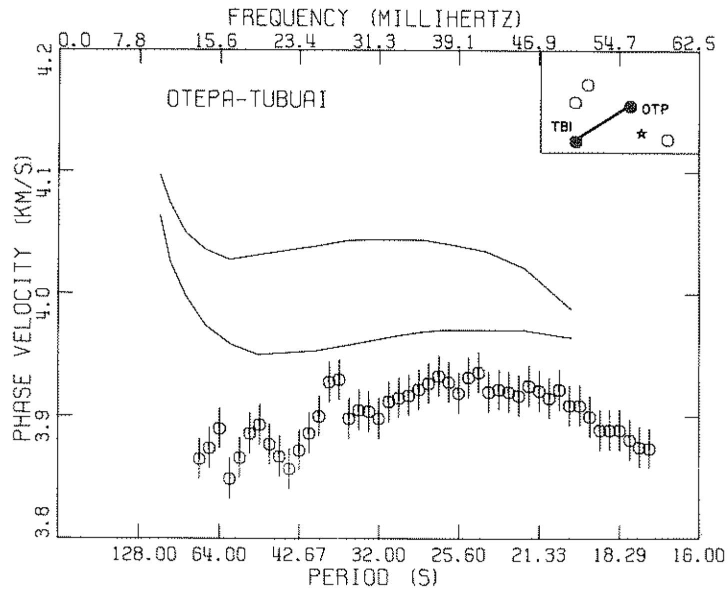


**Figure 5.** Same as Fig. 3 for the path OTP–TBI.

of the wave, we obtain anisotropy coefficients $a(T)$ varying from 1.7 per cent around 17 s to 2.7 per cent at 70 s. Although these figures are comparable to the amount of anisotropy reported by Talandier & Bouchon in their $P_n$ study, they are in excess of the values found by Forsyth (1975; $\leqslant 1$ per cent), and Mitchell & Yu (1980; 0.6 per cent). Also, a number of problems affect the use of equation (2) when trying to fit our data: first, the values along the path TBI–RKT, although fewer, are systematically too low to fit equation (2), once the parameters $a(T)$ are obtained from the two extreme paths. Furthermore, as we will discuss below, the concept of anisotropy being 'frozen' into the plate at the ridge (Francis 1969) predicts an axis of maximum velocity oriented along the direction of spreading at the ridge *where and when* the portion of lithosphere involved was generated. In this area of the Pacific, the plate was generated at the old Pacific–Farallon ridge system (Herron 1972) and the ancient direction of spreading, which was parallel to fracture zones such as the Marquesas and Austral, strikes about 258°. It is, however, impossible to fit our data with such a value of $\phi_{\mathrm{pl}}$, which would necessitate the fastest dispersion to be along the paths RKT–TBI and OTP–TBI, with RKT–TPT slower. This cannot be reconciled with our data. Finally, the interpretation of azimuthal variation in the dispersion as being due to structural heterogeneity needs to be discussed, since local structures selectively affecting our various paths could significantly contribute to the observed variation. In particular, zones of anomalous propagation, involving strong attenuation of body and 20-s surface waves are known to exist in French Polynesia (Talandier & Bouchon 1979; Talandier, unpublished). These probably correspond to magmatic chambers, and are presumably also zones of very low shear velocity. The low phase velocities observed along OTP–TBI could be the result of the existence of such a structure along the path. However, we think that we can discard this possibility, since such anomalies, observed in the Tahiti–Méhétia area and between Tahiti and Rangiroa, are associated with well documented seismicity (Talandier & Kuster 1976). No seismicity is known along the path OTP–TBI, or in its vicinity, although detection capabilities in the area are adequate (Okal *et al.* 1981). More importantly, since the fastest path TPT–RKT lies almost entirely over the Tuamotu island chain, these islands may be an additional source of heterogeneity. It is therefore desirable to measure the contribution of the island structure to the apparent anisotropy, by comparing results from the path TPT–RKT with results involving a path parallel to it, but outside the island chain. It is tempting to use the alignment RKT–PPT, but multipathing is often observed at PPT for teleseismic records following this path. In the next section, we will circumvent this problem by making use of records of a local event, whose location between RKT and PPT is unfavourable to multipathing.

## 3 Local results outside the Tuamotu chain

We use a 1977 event (No. 10 in Table 2), located by the USGS in the vicinity of Mururoa, and well recorded at PPT and RKT. Since this event is given a very shallow depth, its two-lobe dip-slip spectral contribution ($iq_R Q_R^{(1)}$ in the notation of Kanamori & Stewart 1976) will be negligible, as $Q_R^{(1)} = 0$ at the surface, and we can assume that its source phase has no azimuthal dependence. We can then use the one-station method at PPT to study the path epicentre 10–PPT, or the two-station method between PPT and RKT to study the path PPT–epicentre 10–RKT. In this latter case, the inverse of the phase velocity computed is a linear combination of the values along the paths to the two stations:

$$1/C = \frac{1}{\Delta_{\mathrm{PPT}} - \Delta_{\mathrm{RKT}}} \cdot (\Delta_{\mathrm{PPT}}/C_{\mathrm{PPT}} - \Delta_{\mathrm{RKT}}/C_{\mathrm{RKT}}), \tag{3}$$

**Table 4.** Phase velocities obtained from event 10.

| Period (s) | Phase velocity (km s$^{-1}$) (PPT–epicentre 10–RKT) |
|---|---|
| 28.98 | 4.102 |
| 28.44 | 4.073 |
| 27.93 | 4.054 |
| 27.43 | 4.015 |
| 26.48 | 4.054 |
| 25.60 | 4.034 |
| 25.18 | 3.985 |
| 24.77 | 4.025 |
| 24.38 | 4.001 |
| 24.00 | 3.977 |
| 22.93 | 3.997 |
| 22.59 | 3.949 |
| 21.63 | 3.969 |
| 20.76 | 3.965 |
| 20.48 | 3.958 |
| 20.21 | 3.942 |
| 19.95 | 3.921 |
| 19.44 | 3.949 |
| 19.20 | 3.941 |
| 18.73 | 3.946 |
| 18.51 | 3.936 |
| 18.29 | 3.936 |
| 18.07 | 3.931 |
| 17.86 | 3.924 |
| 17.66 | 3.919 |
| 17.45 | 3.895 |
| 17.26 | 3.887 |
| 17.07 | 3.906 |

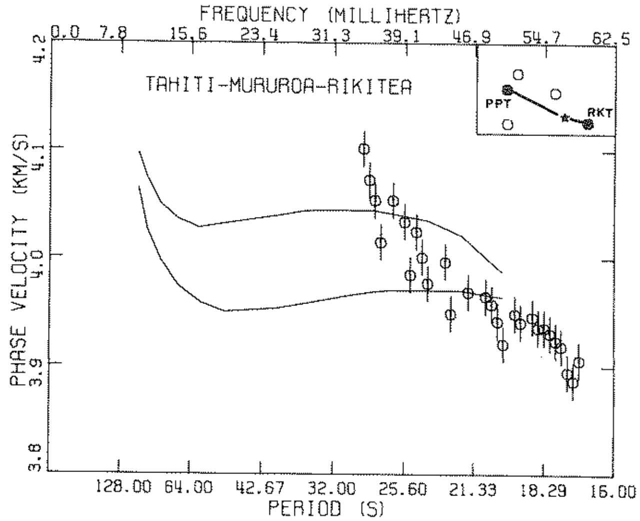


**Figure 6.** Same as Fig. 3 for the path PPT–epicentre 10–RKT.

$C_{\mathrm{PPT}}$ and $C_{\mathrm{RKT}}$ being the velocities along the paths PPT–epicentre 10 and RKT–epicentre 10 respectively, and the $\Delta$s the distances. Such a computation is meaningful only because $\Delta_{\mathrm{PPT}} \gg \Delta_{\mathrm{RKT}}$. The agreement between the two computations fell within uncertainty bars, and justified our assumption on the source phase.

Results are presented in Table 4 and on Fig. 6. The record at PPT, shown on the bottom of Fig. 7, does not exhibit the beating pattern characteristic of multipathing observed for teleseismic events propagating along RKT–PPT. This is probably due to the fact that multipathing at PPT would require too steep an angle of refraction for the anomalous rays venturing into the Tuamotu archipelago from epicentre 10. We interpret our results on Fig. 6 as representative of the dispersion along the azimuth RKT–PPT exclusive of multipathing. Although they are concentrated at the high-frequency end of the spectrum, they indicate beyond doubt that the path PPT–epicentre 10–RKT is slower, by approximately 1 per cent, than the nearly parallel path RKT–TPT. This reduces the anisotropy between RKT–PPT and OTP–TBI to a figure varying from 1 to 2 per cent (on average 1.5 per cent), resulting in an average coefficient *a* of only 1.1 per cent, in much better agreement with earlier reports (Forsyth 1975; Mitchell & Yu 1980). Furthermore, the third path, RKT–TBI, then fits equation (2) within error bounds.

We will now study the influence of the structural lateral heterogeneity between the Tuamotu chain and the nearby ocean, on the propagation of teleseismic Rayleigh waves in the area.

## 4 Multipathing for teleseismic Rayleigh waves along RKT–PPT

Fig. 7 compares the records at TPT and PPT for event 11 (see Table 2), a 1970 shock in southern Chile. Unfortunately, station RKT was not equipped with a long-period instrument at the time. The difference in wave form at the two stations is striking, especially given their very similar epicentral distances and azimuths. A great-circle to TPT from the epicentre travels along the Tuamotu archipelago, whereas a great-circle to PPT misses the archipelago, and travels through the area studied in Section 3. For reference, Fig. 7 also shows that waves reaching PPT from epicentre 10 do not exhibit the same pattern suggestive of beats, present

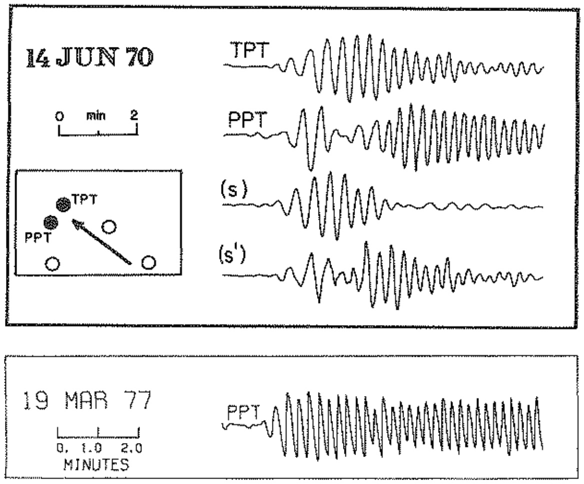


**Figure 7.** Upper box: the top two traces are observed seismograms for event 11 at TPT and PPT. Note the complex waveform at PPT, suggesting beats; (s): synthetic seismogram expected at PPT on the basis of only a difference in dispersion; (s'): synthetic seismogram obtained by letting TPT and (s) interfere (see text). Lower box: observed trace for event 10 at PPT. Note the absence of the beating pattern.

in the 1970 record at PPT but absent from TPT. In view of the faster velocities along the Tuamotu archipelago, we interpret the anomalous waveform at PPT as the result of multipathing: we propose that the signal results from the interference of a direct wave travelling along the great-circle path, and of a second, anomalous, wave, travelling through the faster island chain, and eventually refracted back to the island of Tahiti. Multipathing of surface waves is a common occurrence in laterally heterogeneous structures, and has been reported by Capon (1970) at LASA and Hadley (1978) in southern California. In order to justify this interpretation, we must first rule out the possibility that the difference in waveshape is due merely to the difference in dispersion between the archipelago and the nearby ocean. For this purpose, we construct an 'expected direct' seismogram at PPT by the following procedure: we take the observed record at TPT, and replace the distance travelled along the island chain to TPT by the distance travelled in the adjoining ocean to reach PPT. Namely, at each angular frequency $\omega$, we modify the spectral phase $\phi_{TPT}$ into:

$$\phi_s = \phi_{TPT} + i\omega\left[\frac{\Delta_{TPT}}{C_1(\omega)} - \frac{\Delta_{PPT}}{C_2(\omega)}\right]. \qquad (4)$$

Here, $\Delta_{TPT} = 1700$ km is the total length travelled along the archipelago to TPT, $\Delta_{PPT} = 1615$ km reflects the slightly shorter distance to PPT. $C_1(\omega)$ and $C_2(\omega)$ are the (smoothed) phase velocities given in Figs 3 and 6 for the corresponding paths. We assume identical propagation from the epicentre to the south-eastern end of Polynesia, and identical initial phases, in view of the small difference in azimuth. We then go back to the time domain, and obtain the third trace, labelled (s) on Fig. 7, which represents the expected waveform, if we allow only for a difference in dispersion to explain the different signals observed. It turns out that (s) is even less dispersive than the original signal at TPT, and therefore cannot fit the observed data at PPT. This rules out the possibility that the difference in waveshape is due to the difference in dispersion between the archipelago and the ocean, and strongly suggests the presence of multipathing.

In order to model the multipathing, we would need to know accurately the path of the anomalous wave. Although it is conceivable to use Fermat's principle in this respect, this would require a precise knowledge of the variations of the phase velocity field over distances comparable to the wavelengths involved (typically 80 km). This is clearly beyond our present resolution, and therefore, we choose to simply add the original TPT record (representative of the waveform in the archipelago) to the (s) trace, after time-lagging. In doing so, we are neglecting a large number of phenomena over which we have little if any control, such as a possible phase shift at the refraction point of the anomalous wave, the influence of the propagation between the refraction and PPT, and so on. Also, we assume that the amplitudes of the two interfering waves are identical. Nevertheless, under these very crude simplifications, and for a time delay between the two waves of 27 s, we obtain the fourth trace, labelled (s′) on Fig. 7, which clearly shares a beating pattern with the PPT signal. This crude modelling shows that the observed anomalous waveform at PPT is due to interference between two waves having travelled different paths, the second (anomalous) one taking advantage of the faster dispersion along the nearby Tuamotu archipelago. The remaining discrepancies between the observed PPT record and (s′), as well as the value of the delay (27 s) are probably related to structural details, on a scale much finer than can be envisioned in the present paper.

## 5 Discussion and conclusion

We have obtained experimental data on Rayleigh wave phase velocities along paths criss-crossing French Polynesia at various azimuths. We confirm the adequacy of models obtained independently from regionalizations of paths sampling lithosphere of variable age: our average values fall between Mitchell & Yu's (1980) models 2 and 3, as expected from our inferred lithospheric ages. We also give a direct confirmation of the existence of azimuthal anisotropy in Rayleigh wave dispersion, which had been inferred from difficulties encountered in regionalizations (Forsyth 1975; Yu & Mitchell 1979). However, a substantial part of the apparent anisotropy in our results is due to the presence of the Tuamotu chain along one of our paths of study. This leaves a genuine anisotropy of only about 1.5 per cent, still significantly larger than the uncertainties in our values, and yields an anisotropy coefficient $a \simeq 1.1$ per cent, comparable to the estimates found in the literature, and confirming the anisotropic structure of the lithosphere evidenced in the same area by Talandier & Bouchon through the use of $P_n$ waves. However, at the few periods (around 27 s) for which we have data along both PPT–epicentre 10–RKT and RKT–TBI, it is clear that the former is faster. This suggests that the axis of maximum velocity may be related to the present stress field, rather than to the stress field at the time of creation of the plate at the ridge. This is in apparent contradiction with most of the results from previous studies, which have generally indicated a coincidence between direction of maximum velocity and direction of spreading at the ridge, where and when the plate was formed (Morris, Raitt & Shor 1969; Raitt *et al.* 1969; Keen & Barrett 1971; Yu & Mitchell 1979). The concept of anisotropy being 'frozen' into the plate at its creation was developed by Francis (1969). On the other hand, Nur & Simmons (1969) have shown that anisotropy can develop in a rock as a result of a uniaxial stress being imposed, the direction of maximum velocity being that of the applied (compressional) stress. It is therefore important to discuss critically all the relevant data, to see whether our observations can be reconciled with Francis' (1969) models. Most of the $P_n$ data, concerning the seismic velocity just below the crust, indicate unambiguously a maximum-velocity axis in the direction of the original spreading at the time of plate formation (identified by the axis perpendicular to the magnetic anomalies): Raitt *et al.* (1969) off the coast of California, Morris *et al.* (1969) in the area of Hawaii, and Keen & Barrett (1971) off the western coast of Canada, all reported this fit to be better than 20°, whereas the angle between present spreading and original spreading was in some cases on the order of 45° (although Talandier & Bouchon (1979) claim a maximum $P_n$ velocity in the direction of present spreading in Polynesia, their data cannot resolve the question, since the $P_n$ velocity field may be perturbed by the presence of the Tuamotu archipelago). Thus, $P_n$ does show a definite correlation with the original direction of spreading, but it is probably representative of only the shallowest layers of the mantle (Ave'Lallemant & Carter 1970). Deeper parts of the upper mantle are correctly sampled only by surface waves.

In his extensive study of the Nazca plate, Forsyth (1975) presented evidence of both azimuthal and polarization anisotropy, but the coincidence of the present and original directions of spreading in this plate prevented him from resolving the indeterminacy. Results in the Pacific plate are less coherent: Schlue & Knopoff (1976, 1977) found no statistically significant evidence for azimuthal anisotropy. They concentrate intrinsic polarization anisotropy in the asthenosphere; Yu & Mitchell (1979) and Mitchell & Yu (1980), on the contrary, require anisotropic material only in the lid (as does Forsyth), and observe an azimuthal anisotropy whose axis of maximum velocity (if assumed constant in the entire plate) coincides with the direction of original spreading (265° ± 6° (1979); 268° ± 2° (1980)).

When considering the elastic properties of the lower lithosphere, it is to be kept in mind that over periods of several million years, these layers behave plastically (e.g. McNutt & Menard 1978); independent evidence for a plastic layer in the deep lithosphere has also been given by Chen & Forsyth (1978). The temperature in the deeper lithosphere rises to more than 1000°C, well above the point for syntectonic recrystallization (Ave'Lallemant & Carter 1970). Depending on the temperature and stress conditions, this recrystallization may be of the rotation or migration type (Guillope & Poirier 1979), the latter leading to structural anisotropy on a coarser scale, brought about by the dislocations accompanying the viscous flow involved in the response of the plate to stress (Carter 1976). These phenomena take place over a period of time $\tau = \eta/\mu$, where $\eta$ and $\mu$ are respectively the viscosity and rigidity of the material, and will thus have had ample time to occur since the ridge jump and re-orientation of spreading in the eastern Pacific, about 15 Myr ago (Herron 1972), for any material of viscosity $\eta \leqslant 10^{26}$ poise. This bound is clearly larger than any reasonable estimate of the lithosphere's viscosity (Walcott 1970); this is confirmed by the fact that over shorter periods of time (e.g. 2 Myr), the lower lithosphere in Polynesia behaves plastically (McNutt & Menard 1978). We come to the conclusion that stresses applied in the present plate motion pattern have probably created structural (if not crystalline) anisotropy in the lower lithosphere since the ridge jump, this anisotropy being of course oriented parallel to the stress. Also, the deepest portions of the lid have probably crystallized along the asthenosphere–lithosphere boundary, far away from the ridge, and therefore possibly quite recently, under the new conditions of stress. Any anisotropy frozen into the deepest portion of the plate at that time would then be oriented along the present direction of spreading.

Most of the present seismic observations would then be satisfied by a model involving crystalline anisotropy 'frozen in' from the ridge in the cold, thin layers just below the Moho discontinuity, and structural (or even crystalline) anisotropy due to plate tectonic stresses (and therefore oriented parallel to the present direction of spreading) in the deeper parts of the lithosphere. Anisotropy related to asthenospheric flow (and conceivably oriented in relation to the absolute motion of the plate) could exist in the channel. The only data which would not substantiate the present model is the orientation of Mitchell & Yu's (1980) axis of azimuthal anisotropy; however, our local investigation indicates that Rayleigh wave anisotropy is both larger than, and oriented at least 30° away from, Mitchell & Yu's model, which is based on the simple assumption of a homogeneous Pacific-wide anisotropy. This may suggest an estimate of the lateral variation of this anisotropy in the Pacific.

A physical model for the existence of the anisotropy, derived by Schlue (1977) involves the opening of cracks in the plate, and their filling with liquid magma. If the orientation (or reorientation) of the cracks is governed by the prevailing stress conditions, the resulting anisotropy would be along the present direction of spreading. Schlue's model predicts a much lower anisotropy for Love waves; this, however, cannot be tested at the present time, due to the absence of horizontal records in the area.

In conclusion, our data provide an independent, and positive, check on the validity of regionalization procedures in surface-wave dispersion studies; this would suggest that the assumptions underlying such inversions (such as lateral dependence governed only by the age of the lithosphere, etc.) are justified. The observed azimuthal anisotropy correlates well with the present stress pattern in the plate, which suggests that the lower lithosphere has undergone a plastic relaxation, compatible with the present estimates of its viscosity. Finally, the Tuamotu archipelago is a zone of high velocities, which is responsible for Rayleigh-wave multipathing.

Due to the relatively small size of the dataset in the present study, consisting exclusively of Rayleigh-wave phase velocities, it seems unwarranted to try to invert it into a shear-velocity profile, especially since a model for the relevant lithospheric age is already available (Mitchell & Yu 1980).

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